Crustal Cracks and Frozen Flow in Oceanic Lithosphere Inferred From Electrical Anisotropy

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Abstract Geophysical observations of anisotropy in oceanic lithosphere offer insight into the formation and evolution of tectonic plates. Seismic anisotropy is well studied but electrical anisotropy remains poorly understood, especially in the crust and uppermost mantle. Here we characterize electrical anisotropy in 33 Ma Pacific lithosphere using controlled-source electromagnetic data that are highly sensitive to lithospheric azimuthal anisotropy. Our data reveal that the crust is ~18–36 times more conductive in the paleo mid-ocean ridge direction than the perpendicular paleo-spreading direction, while in the uppermost mantle conductivity is ~29 times higher in the paleo-spreading direction. We propose that the crustal anisotropy results from subvertical porosity created by ridge-parallel normal faulting during extension of the young crust and thermal stress-driven cracking from cooling of mature crust. The magnitude of uppermost mantle anisotropy is consistent with recent experimental results showing strong electrical anisotropy in sheared olivine, suggesting its paleo-spreading orientation results from sub-Moho mantle shearing during plate formation.

Plain Language Summary A major goal in geoscience is to understand the creation and evolution of oceanic lithosphere. To that end, geophysicists study how properties like electrical conductivity vary with direction and depth in the oceanic lithosphere. We call such directional variation “electrical anisotropy.” Since electrical conductivity is particularly sensitive to fluids, certain minerals, and past deformation, the patterns of electrical anisotropy in the crust and mantle provide evidence for how the lithosphere forms and evolves. For the first time, we use an active-source electromagnetic technique to constrain the electrical anisotropy of Pacific oceanic crust and the shallowest portions of the mantle. Our model shows that the oceanic crust and uppermost mantle are highly anisotropic. We interpret the electrical anisotropy in the crust as fluid-filled cracks that parallel the paleo mid-ocean ridge. This suggests that crustal electrical structure begins to form at the mid-ocean ridge and continues to evolve over time through those early weaknesses. If such cracks are also present in other tectonic plates, then the oceanic crust may be a more important reservoir of water than previously thought. Uppermost mantle electrical anisotropy is consistent with strong shear deformation of olivine that freezes into the mantle early in its formation.

1. Introduction

After seeking the signature of seafloor electrical anisotropy in the PEGASUS marine controlled-source electromagnetic (CSEM) data (Constable & Cox, 1996) using half-space models, Everett and Constable (1999) concluded that “Results such as these, based on only five data points, beg the question of verification.” In 1998 while attending the 14th EM Induction Workshop in Sinaia, Romania, a group of scientists developed a plan for verification and later wrote a proposal called Anisotropy and Physics of the Pacific Lithosphere Experiment (APPLE) to fund it. Data collection in 2001 included a 30 km radius circular deep-tow of an electromagnetic (EM) transmitter, with sensitive seafloor electric field receivers at the circle’s center. Behrens (2005) worked on these data, but the modeling code available to him at the time had only advanced to three-layer models with a single anisotropic layer. In this work we bring modern computational tools to the APPLE data set to show that the oceanic crust and uppermost mantle are both azimuthally anisotropic, with a conductive ridge-parallel direction in the crust due to faulting and cracking, and a conductive direction in the mantle consistent with spreading-parallel shearing.

Anisotropy, which denotes a directional variation in a material property, is becoming a standard feature in geophysical characterizations of oceanic plates because it provides a window into lithospheric development.
and evolution. Seismic studies have illuminated and interpreted anisotropy in the lithospheric mantle using its magnitude and direction to infer mantle flow (Gaherty et al., 2004; Kodaira et al., 2014; Mark et al., 2019; Russell et al., 2018; Shearer & Orcutt, 1985; Vanderbeek & Toomey, 2017). Electrical anisotropy, which can arise from a crystal’s intrinsic properties (Yang et al., 2012), mineral-scale fabrics that produce a lattice or crystal-preferred orientation (Karato et al., 2008), or macroscopic features such as fault-enhanced hydration pathways (Key et al., 2012; Naif et al., 2015), offers complementary insight into crust-mantle processes and the history of lithospheric development.

The general isotropic resistivity structure of normal oceanic lithosphere is well known from borehole logging and CSEM studies. Seafloor sediments and upper crustal volcanics have a relatively low resistivity (∼10Ω m), which rises rapidly from ∼10²–10³ Ω m in the dikes to ∼10⁵–10⁶ Ω m or more in the underlying gabbro and upper mantle peridotite (Becker et al., 1982; Constable & Cox, 1996; Cox et al., 1986; Naif et al., 2015). The base of the lithosphere is too deep for present CSEM studies to resolve, but magnetotelluric (MT) and laboratory mineral conductivity studies show that resistivity falls at depths of 50–100 km as temperature increases. Even larger decreases in resistivity are possible in regions containing partial melt or hydrated mantle olivine (Evans et al., 2005; Naif et al., 2013; Pommier, 2013).

EM geophysical data are particularly sensitive to fluids conductive mineral phases, and melt, making them ideal for studying areas influenced by hydration (Key et al., 2012; Naif et al., 2015), regions with anisotropic fabrics (Le Masne & Vasseur, 1981), and areas presumed to contain partial melt (Baba et al., 2006; Evans et al., 2005; Key et al., 2013; Naif et al., 2013). However, relatively few electrical anisotropy studies have been attempted in oceanic settings. Of these, several have employed the low-frequency MT method to study the deep and conductive asthenosphere (Baba et al., 2006; Evans et al., 2005; Naif et al., 2013; Sarafian et al., 2015), and in some cases the resistive lithospheric mantle (Matsuno & Evans, 2017; Sarafian et al., 2015). Authors in these studies estimate electrical anisotropy in the asthenospheric mantle to range in magnitude from 0 (Sarafian et al., 2015) to a factor of 4 (Evans et al., 2005) and have interpreted the anisotropy as having resulted from aligned melt tubes (Naif et al., 2013) or anisotropic hydrogen diffusion in preferentially oriented olivine grains (Evans et al., 2005). The target in this study is the shallow and resistive uppermost lithosphere, which is too cold to permit widespread melting, apart from regions of focused magmatism like hotspot volcanism. Additionally, the storage capacity of water in olivine at uppermost lithospheric mantle pressures and temperatures preclude the prevalence of hydrous olivine (Mierdel et al., 2007). Therefore these mechanisms are unlikely to explain the crust and uppermost mantle anisotropies required by the CSEM data analyzed in this study.

Though MT data have proven useful in expanding our understanding of asthenospheric electrical conductivity, such data lack resolution and sensitivity to anisotropy in the shallow, resistive portions of the uppermost lithosphere (Cox, 1981; see Section 3.2.3). Some EM studies have used the marine CSEM method (Cox et al., 1986; Everett & Constable, 1999; Key et al., 2012; Naif et al., 2015), which can better constrain shallow lithospheric resistivity and anisotropy (Section 3.2). However, most deep ocean CSEM studies have relied on survey geometries that poorly constrained the magnitude and direction of anisotropy (Cox et al., 1986; Everett & Constable, 1999). Thus, our description of electrical anisotropy in the uppermost oceanic lithosphere is inadequately constrained by previous field studies.

Here we present an anisotropic inversion of CSEM data collected over 33 Ma Pacific lithosphere using a survey geometry that is highly sensitive to azimuthal electrical anisotropy in the upper 20 km of the lithosphere. We show that, to fit the observed data, both the crust and uppermost mantle require at least an order of magnitude of anisotropy with orthogonal directions of maximum conductivity.

1.1. Interpreting the Electrical Conductivity Tensor

The most general, mathematical representation of electrical anisotropy takes the form of a symmetric, 3 × 3 conductivity tensor. Here, we restrict our analysis to the simpler, triaxial tensor, $\hat{\sigma}$, which is compatible with the anisotropy observed in the APPLE data:

$$\hat{\sigma} = \begin{pmatrix} \sigma_x & & \\ & \sigma_y & \\ & & \sigma_z \end{pmatrix},$$

where $\sigma_x$, $\sigma_y$, and $\sigma_z$ refer to conductivity along the principal axes, $x$, $y$, and $z$. In the case of full triaxial anisotropy, $\sigma_x \neq \sigma_y \neq \sigma_z$, whereas isotropic media obey $\sigma_x = \sigma_y = \sigma_z$. For triaxial anisotropy, Ohm’s law
Figure 1. Physical realizations of transverse isotropy (adapted from; Everett & Constable, 1999, Figure 3). In this case, $\sigma_x = \sigma_z = \sigma_0$ and $\sigma_y = \sigma_1$ with $\sigma_0 \neq \sigma_1$. (a) The physical realization of $\sigma_0 > \sigma_1$ is conductive vertical sheets in the $x$-$z$ plane. (b) $\sigma_0 < \sigma_1$ can be realized by conductive rods in the $y$ direction. The other two types of transverse isotropy are obtained upon rotations of the conductivity tensor and corresponding sheet/rod structure.

can be written as

$$
\begin{pmatrix}
J_x \\
J_y \\
J_z
\end{pmatrix} =
\begin{pmatrix}
\sigma_x & \sigma_y & \sigma_z \\
\sigma_y & \sigma_0 & \sigma_z \\
\sigma_z & \sigma_z & \sigma_0
\end{pmatrix}
\begin{pmatrix}
E_x \\
E_y \\
E_z
\end{pmatrix},
$$

which shows that anisotropic conductivity imparts a directionally dependent scaling on the electric field components $(E_x, E_y, E_z)$. The resulting current densities $(J_x, J_y, J_z)$ can be either preferentially enhanced or reduced in each direction, depending on the nature of the anisotropy.

As we will show, the APPLE data can be fit with a transversely isotropic model in which only one of the three conductivity components is distinct from the other two. It is relevant to discuss the physical realizations of such anisotropy. Consider the case in which $\sigma_x = \sigma_z = \sigma_0$ and $\sigma_y = \sigma_1$ so that

$$
\sigma =
\begin{pmatrix}
\sigma_0 & \sigma_1 \\
\sigma_1 & \sigma_0
\end{pmatrix}
$$

Everett and Constable (1999) show that two different geological media can describe such anisotropy depending on the relative conductivities of $\sigma_0$ and $\sigma_1$. The case where $\sigma_0 > \sigma_1$ corresponds to conductive sheets oriented in the $x$-$z$ plane and embedded within a more resistive background (Figure 1a). Alternatively, $\sigma_0 < \sigma_1$ corresponds to conductive lineations (rods) aligned in the $y$ direction and embedded in a more resistive background (Figure 1b). Two additional types of transverse isotropy follow naturally upon rotation of the conductivity tensor and similarly correspond to conductive sheets and rods.

2. Data Acquisition

The APPLE experiment was centered about 1,000 km west of California, USA, on 33 Ma old lithosphere that formed at the fast-spreading East Pacific Rise (EPR). The region was deemed logistically convenient for studying anisotropy since it is far from potential lateral heterogeneities at nearby plate boundaries, has only about 100 m sediment cover (Divins, 2003) to attenuate the CSEM signal, and has low bathymetric relief of up to $\sim$500 m. The paleo-spreading direction for this region was approximately due east-west and the full paleo-spreading rate was $\sim$13 cm/yr (Müller et al., 2008). North-south trending escarpments in high-resolution bathymetry indicate normal faults consistent with abyssal hills that form proximal to mid-ocean ridges (Ryan et al., 2009; Figure 2). We collected CSEM data by deep towing a horizontal electric dipole (HED) source $\sim$100 m above the seafloor while it transmitted a square-wave current with a fundamental frequency of 4 Hz (Behrens, 2005). The survey included a primary 30 km radius circular tow around orthogonal pairs of highly sensitive, long-wire electric field HED receivers with 177 m dipoles (LEMs in Figure 2; Webb et al., 1985). The purpose of this large radius circular tow was to measure azimuthal anisotropy. A storm passed through the survey area that temporarily restricted certain tow headings and prompted the unplanned acquisition of a 15 km radius semicircular tow around receiver Quail. Receiver
Quail is a conventional ocean bottom electric field sensor, known as a Mk I instrument as described in Constable (2013), with 10 m dipoles. Because the signal-to-noise ratio of the receivers is proportional to dipole length, conventional receivers like Quail are much less sensitive than the LEMs (Webb et al., 1985). Long-offset data collected during a radial tow ranging from 14 to 70 km provide additional constraints on the depth dependence of conductivity.

Vector amplitude and phase data for the LEM and Quail receivers were obtained at the transmitter’s 4 Hz fundamental harmonic using 2-min long sections of the raw time series data. These were then averaged to create longer stacks equivalent to 20, 60, and 150 min for the Quail and LEM data, depending on the offset between the receiver and transmitter. Uncertainties in the final stacked data were estimated using the variances of the stack residuals. To account for the uncertainty in the navigation of the source dipole, which is most significant at short offsets, the error floor in uncertainty for data at offsets less than 5 km was set to 25%, while the error floor for data at offsets greater than 5 km was lowered to 10%. Further details of the processing analysis are given in Behrens (2005).

The vector data can be inverted using either the Cartesian field components or electromagnetic polarization ellipse parameters (Smith & Ward, 1974). Here we use polarization ellipse parameters in part because they are visually useful in confirming and evaluating azimuthal anisotropy. For a fixed transmitter-receiver geometry at a given time, each receiver will record a particular vector horizontal electric field. Over the time period of a sinusoidal transmission, the orientation and magnitude of the field vector will oscillate such that its head traces an ellipse (Figure 3). This polarization ellipse is described by a major and minor axis, which we halve to obtain $P_{\text{max}}$ and $P_{\text{min}}$, respectively. The mathematical representations for $P_{\text{max}}$ and $P_{\text{min}}$ are

$$P_{\text{max}} = |E_x| e^{i \Delta \phi} \cos \alpha + |E_y| \sin \alpha$$

$$P_{\text{min}} = |E_x| e^{i \Delta \phi} \sin \alpha - |E_y| \cos \alpha$$
Figure 4. Fit of preferred resistivity model to the data. Symbols show the data and lines show model responses. Dashed model response lines show the fit of the preferred model allowing for lateral, east-west variations in sediment conductivity as described in Section 3.4; that resistivity model is shown in Figure 13. Error bars are 1 standard deviation. (a) Angular dependence of $P_{\text{max}}$ and $P_{\text{min}}$ for the circular tow data recorded by the LEM receivers shown as a function of the transmitter to receiver azimuth angle, $\theta$. The overlain vertical and horizontal lines indicate the paleo-ridge-parallel and paleo-spreading directions, respectively. (b) Radial data recorded by the LEM and Quail receivers. The offset where the range overlaps is due to the different transmitter and receiver geometries. (c) Quail data from the semicircular tow.

where $E_x$ and $E_y$ are the frequency domain orthogonal measurements of the horizontal electric field, and $\Delta \phi$ is given by

$$\Delta \phi = \phi_x - \phi_y$$

where $\phi_x$ and $\phi_y$ are the absolute phases of the x and y electric fields, respectively. Angle $\alpha$ is given by

$$\tan(2\alpha) = \frac{2|E_x||E_y| \cos \Delta \phi}{|E_x|^2 - |E_y|^2}. \quad (6)$$

The orientation and shape of the polarization ellipse depend on the transmission frequency, the transmitter-receiver geometry, and the substrate’s conductivity. Here we focus on data obtained at the 4 Hz fundamental frequency of the transmitter since it has a much larger signal-to-noise ratio compared to the higher-frequency harmonics. Thus the $P_{\text{max}}$ and $P_{\text{min}}$ APPLE data are functions of the relative positions of the transmitter and receiver and the conductivity of the underlying lithosphere.
Figure 5. CSEM sensitivity study for crustal and uppermost mantle anisotropy showing forward responses as a function of receiver-transmitter azimuth, $\theta$, for simplified layered models at a receiver-transmitter range of 30 km. Layer resistivities are shown in rectangles next to each response. Note that the positive $x$ direction is along azimuth $0^\circ$, while the positive $y$ direction is along $90^\circ$. (a) In the isotropic case, where neither the crust nor uppermost mantle is anisotropic, $P_{\text{max}}$ is circular and $P_{\text{min}}$ is zero for all azimuths. (b) Upper layer (crust) is anisotropic with the conductivity 30 times greater in the $x$-$z$ plane than the $y$ direction. This anisotropy can result from conductive vertical sheets (Section 1.1; Everett & Constable, 1999). (c) Lower layer (uppermost mantle) is anisotropic with conductivity 30 times greater in the $y$ direction than the $x$-$z$ plane. This anisotropy can result from conductive, $y$-oriented lineations (Section 1.1; Everett & Constable, 1999). For the anisotropic models in (b) and (c), $P_{\text{max}}$ is no longer circular and has a predominantly $2\theta$ pattern, while $P_{\text{min}}$ is nonzero and displays a dominant $4\theta$ pattern. For the anisotropic models, the responses are symmetric about the two anisotropy axes ($x$ and $y$).
Table 1
Resistivity Model Parameters

<table>
<thead>
<tr>
<th>Layer</th>
<th>Description</th>
<th>Depth (km)</th>
<th>Starting resistivity $\rho_x, \rho_z \mid \rho_y$ ((\Omega) m)</th>
<th>Final resistivity $\rho_x, \rho_z \mid \rho_y$ ((\Omega) m)</th>
<th>Anisotropy penalty weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Air half-space</td>
<td>10</td>
<td>$10^{13}$</td>
<td>10^{13}</td>
<td>Fixed</td>
</tr>
<tr>
<td>2</td>
<td>Sea</td>
<td>0–4.4</td>
<td>0.3</td>
<td>0.3</td>
<td>Fixed</td>
</tr>
<tr>
<td>3</td>
<td>Sediments</td>
<td>4.4–4.5</td>
<td>75</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>4</td>
<td>Volcanics</td>
<td>4.5–5</td>
<td>200</td>
<td>140</td>
<td>30</td>
</tr>
<tr>
<td>5</td>
<td>Dikes</td>
<td>5–7</td>
<td>135</td>
<td>4,000</td>
<td>259</td>
</tr>
<tr>
<td>6</td>
<td>Gabbros</td>
<td>7–11</td>
<td>800</td>
<td>19,500</td>
<td>1,157</td>
</tr>
<tr>
<td>7</td>
<td>Uppermost mantle</td>
<td>11–20</td>
<td>57,000</td>
<td>2,400</td>
<td>80,618</td>
</tr>
<tr>
<td>8</td>
<td>20+</td>
<td>6,000</td>
<td>13,450</td>
<td>10,770</td>
<td>10,821</td>
</tr>
</tbody>
</table>

Note. The starting model was based on the convergence of a LEM-only data inversion, whose starting model was isotropic. A higher anisotropic penalty was imposed on layer 8 because our data lose sensitivity at around 20 km. See Key (2016) for the details of the anisotropy penalty weight.

Equation (4) shows that the $P_{\text{max}}$ and $P_{\text{min}}$ data are independent of the phase offset between the transmitter and receiver, and depend, rather, on the phase difference, $\Delta \phi$, between the two orthogonal receiver components, which were timed by the same clock (Quail) or clocks synchronized to millisecond accuracy (LEMs). Because they depend only on the vector component phase difference, the polarization ellipse parameters are insensitive to the absolute phase of the transmitter, which was not well constrained in the APPLE experiment since it was a relatively early application of marine CSEM.

Given the circular transmitter tow geometry around the LEMs and Quail (Figure 2), we took the vector horizontal electric fields for each data stack and computed polarization ellipse parameters at evenly spaced transmitter locations along the perimeter of the circular and semicircular tows. We show $P_{\text{max}}$ and $P_{\text{min}}$ for all stack windows as a function of transmitter-to-receiver geographic angle $\theta$ in polar plots (Figures 4a and 4c). These plots are useful since they have characteristic and easy to interpret shapes in the presence or absence of azimuthal anisotropy. Over an isotropic, one-dimensional subsurface, $P_{\text{min}}$ will remain zero as the transmitter circumnavigates the receivers because the transmitter and receiver will always have a broadside geometry (or purely azimuthal mode in dipole terminology) that imparts a linear horizontal polarization to the electric field; further, the $P_{\text{max}}$ component remains constant for constant range (Figure 5a). Conversely, azimuthal anisotropy can create nonzero $P_{\text{min}}$ data as the anisotropic components of the conductivity tensor give rise to preferential electric current flow in the anisotropy direction, leading to differences in electromagnetic diffusion and attenuation along the anisotropy axes. This leads to symmetric azimuthal variations in both $P_{\text{max}}$ and $P_{\text{min}}$, as shown for hypothetical crust and uppermost mantle anisotropies in Figures 5b and 5c. Laterally heterogeneous structure could also impart a nonzero $P_{\text{min}}$ component, but this would likely be evident as asymmetry in the polar plots, rather than the symmetric behavior seen in this figure.

It is clear that the observed LEM circular tow data, shown in Figure 4, indicate an azimuthally anisotropic seafloor. In particular, there is a strong $\cos(2\theta)$ pattern in $P_{\text{max}}$ exhibited as a north-south elongation and an east-west dimpling of the field magnitudes, and there is a $\cos(4\theta)$ pattern in $P_{\text{min}}$ expressed as a cloverleaf shape. These symmetric patterns dominate the variation in the LEM circular tow data, while asymmetric components are of second order, suggesting that any lateral conductivity heterogeneities in the survey region are relatively small in magnitude. These relatively small asymmetries are likely due to subtle variations in sediment thickness as discussed in Section 3.4.

3. Modeling APPLE Anisotropy
3.1. Preferred Inversion Model
Based on approximate crustal boundaries of sediments, volcanics, dikes, and gabbros of the eastern Pacific Plate, we constructed a simple layered resistivity model (Table 1) and performed a nonlinear anisotropic
Figure 6. Preferred resistivity model with 70% and 95% confidence intervals derived from a linearized uncertainty analysis (resistivity is the reciprocal of conductivity). Orange, solid, and blue, dashed lines show resistivity in the paleo-ridge-parallel ($x$-$z$) plane and paleo-spreading ($y$) direction, respectively, for this study. Overlain dashed pink and green lines are resistivities for hydrated bending fault plane parallel and perpendicular directions, respectively, from the Middle America Trench (Key et al., 2012). Experimentally measured conductivity for sheared olivine samples (Pommier et al., 2018; shear strain $= 1.3$) extrapolated to lower temperatures for the shear parallel direction (purple rectangle) and shear plane perpendicular direction (gold rectangle). Because the model space of this inversion is bounded ($10^{-1} \, \Omega\, m \leq \rho \leq 10^{5} \, \Omega\, m$) to avoid unrealistic solutions, the upper confidence limits in the more resistive direction lie outside the plotting range from 7–20 km.

inversion of the circular and radial tow data using an adaptive finite element-based inversion code (Key, 2016). We neglect bathymetry in the layered model because the coupling of the HED source to the seafloor depends primarily on the altitude of the source above the seafloor and not on the absolute bathymetry of the region. This is especially true for the APPLE data where long offsets ($\sim 30$ km) between the source and receivers mean that the dominant mode of energy propagation is through the crust and mantle and not directly through the seawater. Imposing realistic seafloor topography beneath the source and receiver resulted in less than 4% change in the model response compared to a flat seafloor, which is below the 10% error floor imposed on our data.

We parameterized the model for azimuthal anisotropy such that the resistivity was the same in the vertical plane oriented in the north-south direction (the $x$-$z$ plane) and the resistivity in the east-west direction ($y$) was different (sometimes referred to as transversely isotropic $y$, or TIY anisotropy). This choice of parameterization was supported by the north-south, east-west symmetry observed in the LEM circular tow data (Figures 2 and 4a), as well as forward model studies (Section 3.2), and it is aligned with the paleo-spreading ($y$) direction. Although vertical anisotropy is commonly observed in thick sediment layers of continental shelf environments, where beds of alternating grain size can produce a macroscopic anisotropy with the conductive plane parallel to bedding (Key, 2012), we do not expect such anisotropy to be prominent in the relatively thin and uniform abyssal clays of the APPLE region. Thus, here our focus is on azimuthal anisotropy.

This single mathematical description of TIY anisotropy can be interpreted as two different types of geological structures, depending on whether the east-west ($y$) direction is more conductive or more resistive than the north-south-vertical ($x$-$z$) plane. If the east-west direction is more resistive (less conductive) than the north-south-vertical plane conductivity, then the model describes conductive vertical sheets with a
north-south strike direction. Alternatively, if the east-west direction is more conductive (less resistive),
then the model corresponds to conductive lineations with an east-west strike (Everett & Constable, 1999;
Section 1.1 and Figure 1).

To avoid unrealistically high or low resistivities being produced by the inversion, the model space is bounded
by a priori constraints spanning from a resistivity typical for porous sediments (10⁻¹ Ω m) to that typical of
highly resistive, crystalline lithosphere (10⁵ Ω m; Palacky, 1988). These bounds are implemented using the
nonlinear band-pass parameter transform method (Key, 2016). We also imposed regularization that penal-
ized sharp resistivity jumps across layer boundaries and anisotropy within layers. In inverting the azimuthal
and radial data, we used a standard regularized inversion method that minimized the norm:

\[ U = \mu^{-1} ||W(d - F(m))||^2 + ||Rm||^2 \]  

(7)

where \( d \) is the data vector, \( F(m) \) is the forward response for model parameters \( m = [m_{x}, m_{y}] \), where the
elements of \( m \) are scaled as \( \log_{10}(\text{resistivity}) \), and \( W \) is a diagonal matrix of inverse standard errors on the
data. \( Rm \) is a measure of model roughness:

\[ ||Rm||^2 = ||R_{m_{x,y}}||^2 + ||R_{m_{y}}||^2 + \beta ||m_{x,y} - m_{y}|| \]  

(8)

where \( R \) is an operator approximating the spatial gradient of the model parameters. Here we use units of
\( \log_{10}(\Omega m) \) for the model parameters, and the subscripts correspond to the anisotropic components. The first
two terms penalize sharp resistivity jumps across layer boundaries, while the last term penalizes the log of
the anisotropy ratio within layers so that only anisotropy required to fit the data is present in the final model
(Key, 2016). The parameter \( \mu \) is automatically updated in each iteration to balance the data misfit and model
roughness terms. We imposed a 100 times stronger anisotropy penalty, \( \beta \), at depths exceeding 20 km since
our data lose sensitivity below this depth (Table 1).

Our preferred model (Figure 6) fits the data with an RMS misfit of ~1.36. The model response qualitatively
fits the north-south elongations and east-west dimple in LEM \( P_{\text{max}} \) data as well as the cloverleaf pattern
of the \( P_{\text{min}} \) data, along with the amplitudes of the radial tows (Figure 4). The upper crust is anisotropic
with \( \rho_{\text{high}}/\rho_{\text{low}} \sim 18 \) while the lower crust has \( \rho_{\text{high}}/\rho_{\text{low}} \sim 36 \), both with greater conductivity parallel to the
paleo-ridge vertical plane. The uppermost mantle is also anisotropic with \( \rho_{\text{high}}/\rho_{\text{low}} \sim 29 \), but the conduc-
tivity is greater in the orthogonal, paleo-spread direction. Both the resistive and conductive directions in
each layer above 20 km follow the trend of increasing resistivity described from borehole logging and other
CSEM studies (Becker et al., 1982; Constable & Cox, 1996; Cox et al., 1986; Naif et al., 2015).

Before examining the robustness of this model and discussing its geological implications, it is useful to
use some simple forward models to show how it is possible that our data set can even be sensitive to both
north-south striking, conductive vertical sheets, and east-west conductive lineations.

### 3.2. Sensitivity of EM Methods to Lithospheric Azimuthal Anisotropy

Here we expand upon the simple model study shown in Figure 5 to examine how azimuthal anisotropy
manifests itself in azimuthal tow data and to explain the sensitivity of the experimental layout used in APPLE
for interrogating azimuthal electrical anisotropy. We consider CSEM forward responses of a layered resistiv-
ity model based on our preferred inversion model, with a two-layer crust of relatively conductive sediment
underlain by resistive volcanics, dikes, and gabbros, which are further underlain by the uppermost man-
tle. We consider three versions of this model: an isotropic lithosphere (all layers isotropic), an anisotropic
crustal layer atop an isotropic lithospheric mantle, and an anisotropic uppermost lithospheric mantle layer
embedded in an otherwise isotropic lithosphere. The anisotropy ratio is 30 in both of the latter cases, but
the direction (plane) of maximum conductivity is chosen to match our preferred inversion model, with a two-layer crust of relatively conductive sediment
underlain by resistive volcanics, dikes, and gabbros, which are further underlain by the uppermost man-
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use some simple forward models to show how it is possible that our data set can even be sensitive to both
north-south striking, conductive vertical sheets, and east-west conductive lineations.
The maximum amplitude of $P_{\text{max}}$ is greater for crustal anisotropy of this magnitude than it is for uppermost mantle anisotropy, but the opposite is true of $P_{\text{min}}$. In summary, the most obvious feature of crustal anisotropy of this type is the east-west dimpling in $P_{\text{max}}$ (Figure 5b), and for uppermost mantle anisotropy it is the prominent cloverleaf in $P_{\text{min}}$ (Figure 5c). Since both of these features are present in the APPLE data (Figure 4a), we have some confidence that we are observing both types of anisotropy, albeit the specific details of how these two signatures couple when both are present is undoubtedly more complicated. Still, forward modeling a lithosphere with both crustal and uppermost mantle anisotropy confirms the appearance of the $P_{\text{max}}$ dimple and prominent $P_{\text{min}}$ cloverleaf (Figure 7).

Both the crust and uppermost mantle anisotropic responses illustrate the paradox of anisotropy, which is an amplification of the electric field in the direction of maximum conductivity above what would be predicted for an isotropic half-space with the lower of the two conductivities (the more resistive value; Everett & Constable, 1999).

### 3.2.1. Effects of Anisotropy Ratio

Since this is the first study of its kind, we seek to build some intuition for how different variables may present in $P_{\text{max}}$ and $P_{\text{min}}$ data plots for crust-only and uppermost mantle-only azimuthal anisotropy. Figure 8 summarizes the effects of varying the magnitude of TIY anisotropy on $P_{\text{max}}$ and $P_{\text{min}}$ in either the crust or uppermost mantle. Each forward response was generated for a model similar to those of Figure 5 (Section 3.2). For the anisotropic layer, the resistivity of the conductive direction (or plane) was held fixed in each model and that of the resistive direction (or plane) was set equal to the product of the resistivity of the conductive direction and the magnitude of anisotropy. As the magnitude of anisotropy increases, the maxima of both $P_{\text{max}}$ and $P_{\text{min}}$ increase because the resistivity of the resistive direction increases, and thus there is less inductive attenuation of the fields. The rate of increase, however, tends to zero as the magnitude of anisotropy increases, making it more difficult to resolve the magnitude of anisotropy for large anisotropy ratios ($>10$ times). This is a consequence of the saturation of the CSEM electric field response in the resistive direction, that is, when the resistivity in that direction is high enough to no longer significantly attenuate the field over the 30 km transmitter-receiver range.

In all cases modeled, the underlying equations describing the azimuthal dependence of $P_{\text{max}}$ must include a dominant $\cos(2\theta)$ term and that of $P_{\text{min}}$ a dominant $\cos(4\theta)$ term, just as in the APPLE data from the LEM receiver (Figure 4a). Increasing the magnitude of anisotropy has the effect of creating dimples in $P_{\text{max}}$ and increasing the complexity of $P_{\text{min}}$ for crustal anisotropy, or at least the degree to which this complexity is measurable. The dimpling of $P_{\text{max}}$ at $\geq 5$ times anisotropy appears to coincide with the introduction of higher-order terms into the equation describing $P_{\text{max}}$, and, more recognizably, $P_{\text{min}}$. This effect is observed...
3.2.2. Effects of Bulk Resistivity

Figure 9 shows the effects of changing the bulk resistivity of the anisotropic layer while holding the anisotropy ratio fixed.

The resistivity of the conductive direction controls the eccentricity and dimpling of $P_{\text{max}}$, with increased eccentricity and a more pronounced dimple at lower resistivities, likely due to the directionally dependent increase in attenuation of the electric field as the resistivity decreases. The bulk resistivity seems to have a more dramatic effect on the shape of $P_{\text{max}}$ than the anisotropy ratio. It is relevant to note that the dimple does not appear in $P_{\text{max}}$ for the mantle cases modeled here. Intriguingly, the maxima of $P_{\text{max}}$ behave distinctly in the crustal and uppermost mantle anisotropy cases here examined. For crustal vertical sheet anisotropy, decreasing the absolute resistivity of the anisotropic directions increases the maximum of $P_{\text{max}}$. The opposite is true for uppermost mantle horizontal rod anisotropy where decreasing the bulk resistivity decreases the maximum of $P_{\text{max}}$.

The maxima of $P_{\text{min}}$ behave similarly for both the crust and uppermost mantle anisotropy models—decreasing the bulk resistivity of the anisotropic layer increases the overall maximum value attained by $P_{\text{min}}$. It is also interesting to note that, at fixed anisotropy ratio, the relative maxima of $P_{\text{min}}$ are located at the same angles (this was not the case in Section 3.2.1; Figure 8).

3.2.3. Could MT Data Resolve Lithospheric Anisotropy?

MT data have been used to image asthenospheric electrical anisotropies with magnitudes of up to 4 times (Baba et al., 2006; Evans et al., 2005; Naif et al., 2013). Such data are relatively insensitive to lithospheric azimuthal anisotropy. Figure 10 shows the 1-D MT responses for the same models considered in Figure 5.
Figure 9. Forward responses of simple TIY anisotropy models of vertical sheets in the crust and horizontal rods in the uppermost mantle that demonstrate the effects of changing the bulk resistivity on $P_{\text{max}}$ and $P_{\text{min}}$ while holding the anisotropy ratio fixed. The models are outlined in the first column along with their respective physical realizations (Section 1.1), and the forward responses are all other panels. Crustal (uppermost mantle) anisotropy is displayed in the top (bottom) two rows, and the bulk resistivity decreases from left to right in each row. The amplitude labels in the first response column apply to the entire row. Note that there is an order of magnitude difference in the $P_{\text{max}}$ and $P_{\text{min}}$ amplitudes.

The maximum $\sim 20\%$ difference between the parallel and perpendicular responses for the crustal anisotropy occurs at shorter periods of around 10–100 s, which cannot be well-resolved by deep-water, fluxgate magnetometer, marine MT instruments due to attenuation of the source field by the conductive ocean. The uppermost mantle anisotropy creates an even smaller and more difficult to detect split in the MT responses. The crustal and uppermost mantle anisotropy signals are well below the 5–20% data error levels used for the long-period data interpreted as lacking lithospheric anisotropy in Sarafian et al. (2015) and Matsuno and Evans (2017). Broadband MT systems (Naif et al., 2013) capable of measuring periods as short as 20 s in deep water are better suited for anisotropy studies, but they would still have difficulty resolving the...
Figure 10. Forward model study showing that marine MT data are essentially insensitive to azimuthal lithospheric anisotropy of a similar magnitude to our preferred model. These forward models are the same as those used for the CSEM sensitivity analysis shown in Figure 5. (top, middle, and bottom panels) The physical realizations of the anisotropy are shown in each panel for reference (Section 1.1).

maximum signal level of anisotropy found at periods less than 100 s. Additionally, marine MT data are subject to the long ranging effects of the large conductivity contrasts at coastlines as well as regional bathymetry. Such effects can generate an apparent anisotropy in MT responses that is an order of magnitude larger than we have modeled here even if the underlying lithosphere is isotropic and one dimensional (Heinson & Constable, 1992; Key & Constable, 2011; Wang et al., 2019). Given that both broadband and long-period MT data are insensitive to lithospheric azimuthal anisotropy, it is unsurprising that the crust and uppermost mantle anisotropy required by the APPLE CSEM data have not been observed in deep ocean MT surveys.
Figure 11. Uncertainty in the preferred model resistivities. The root-mean-square (RMS) misfits obtained by local grid searches of the model space are shown as colored surfaces. The solid white line is the RMS misfit = 1.5, contour and white stars show the location of the preferred model resistivities for each layer. The dashed white-black line shows where the layer is isotropic. The physical realization of models above the line of isotropy, where $\rho_x, \rho_z < \rho_y$ ($\sigma_x, \sigma_z > \sigma_y$), are the conductive vertical sheets described in Section 1.1. Similarly, conductive, $y$-oriented rods could produce the models below the line of isotropy, where $\rho_x, \rho_z > \rho_y$ ($\sigma_x, \sigma_z < \sigma_y$). The panel title numbers refer to specific layers listed in Table 1. Grid searches over Layers 5 and 6 illustrate the saturation of the CSEM electric field response in the resistive direction.

3.3. Robustness of Preferred Model Anisotropy

To determine the robustness of our preferred model, we estimated the model covariance using a linearized uncertainty analysis to calculate 70% and 95% confidence intervals on the resistivity of each layer. Based on this analysis, the crust and uppermost mantle require anisotropy at the 95% confidence limit (Figure 6). Due to a saturation effect for CSEM responses for highly resistive layers, the upper 70% and 95% confidence limits for the lower crust and the upper 95% confidence limit for the uppermost mantle extend outside the resistivity bounds of our inversion and thus are not shown in Figure 6.

To further assess the necessity for anisotropy in our model and to aid in visualizing the uncertainty of the anisotropy required by our data, we performed a grid search over the anisotropic resistivity components in each layer (Figure 11). To generate a particular misfit surface, the $x, z$, and $y$ resistivities of a given layer were simultaneously varied from $10^6$–$10^7$ $\Omega$ m using a $71 \times 71$ grid, while the other layer resistivities were held fixed at the values of the preferred model. As with the linearized estimate, this grid search suggests that anisotropy is required by the data for the entire crust and the uppermost mantle. It also illustrates the saturation of the CSEM electric field response in the resistive direction, which was similarly noted in the model studies of Section 3.2.1 (Figure 8). Because of this saturation, our data can give a lower bound on the resistivity of each layer, but not necessarily an upper resistivity bound, which here we take to be given by our a priori constraints.
Figure 12. Effect of sediment thickness on $P_{\text{max}}$ and $P_{\text{min}}$. The $P_{\text{max}}$ and $P_{\text{min}}$ forward responses were generated for a model with 50 m of sediment (solid lines) and 150 m of sediment (dashed lines). All other layers have the preferred model resistivities (Figure 6; Table 1). The forward responses assume a perfectly circular, 30 km radius transmitter tow at a constant altitude. Uniformly decreasing the sediment thickness will increase the amplitude of the response (solid lines). Conversely, increasing the sediment thickness decreases the amplitude of the response (dashed lines).

3.4. Lateral Variations in Sediment Thickness

Comparing the azimuthal mode responses of our preferred model with the data reveals certain regions of poorer than average fit (Figures 4a and 4c). In particular, the azimuthal responses underestimate the LEM $P_{\text{max}}$ data from $\sim$110–160$^\circ$ and overestimate the Quail $P_{\text{max}}$ data between $\sim$210–240$^\circ$.

Although the APPLE location was chosen for minimal bathymetry, there is still evidence for abyssal hills in the survey region (Figure 2). Thus, a plausible explanation for these small, second-order data asymmetries is lateral variation in sediment thickness within the survey region due to underlying abyssal hill fabric. In model studies, we find that the sediment thickness acts to scale the magnitudes of $P_{\text{max}}$ and $P_{\text{min}}$ but not their overall azimuthal shapes. Because of attenuation due to its high conductivity relative to underlying mafic volcanics, increasing the sediment thickness decreases the amplitude of the EM responses and vice versa (Figure 12). Hence, CSEM data of this type appear to be sensitive to the total sediment vertical conductance (i.e., conductivity-thickness product), and therefore, either the conductivity or the thickness of the sediment layer could be varied to account for the small second-order features in the data. Thus, upon obtaining our preferred model, which explains the first-order effects of azimuthal anisotropy, we attempted to better fit the observed second-order asymmetries by inverting for east-west variations in conductivity of the uppermost “sediment” layer of our model. We chose to invert for sediment conductivity since our modeling code easily supports that. However, given the conductivity and thickness trade-off mentioned above, and given that deep ocean sediments are typically laterally uniform in conductivity (e.g., Naif et al., 2015), the minor lateral variations in conductivity found by this approach are more likely due to sediment thickness variations from the abyssal hill topography.

To create this model, we fixed the preferred model resistivities for all but the 100 m sediment layer. Within the sediment layer, we created 1 km (east-west) × 10 m (vertical) cells and inverted for the isotropic resistivity of each cell (Figure 13). As this is a 2-D modeling code, cells necessarily extend infinitely in the north-south direction. Cell resistivities converged to values ranging from 1 to $\sim$20 $\Omega$m, or $\sim$0.05 to 1 S/m. If we assumed that the sediments have a uniform conductivity of 1 S/m, as is typical for abyssal plain sediments (e.g., Naif et al., 2015), this would imply a reasonable variation in sediment thickness of $\sim$5–100 m in the survey.
In the uppermost mantle, resistivity is lower in Section 1.1. This is consistent with crustal cracks filled with free water. In the crust, resistivity is lower in Section 1.1. This suggests a shear-induced uppermost mantle anisotropy along the paleo-ridge parallel, which is similar to the conductive, ∼29 times lower in the Ma EPR-derived Cocos crust, which indicates high crustal permeability (Manning et al., 2000), which created ridge-parallel zones of weakness. Such weaknesses promoted deep cracking and openings that allowed for seawater infiltration as the plate aged.

Geologic evidence for this argument lies in analysis of crustal sections from the Hess Deep Rift, a fast-spreading center, which point to dike-parallel fault widening over time (Varga et al., 2004) and reactivation of sealed faults through incremental slip (Hayman & Karson, 2007). Deepening of grabens with distance from the ridge axis has also been observed at the 9°50′N section of the EPR (Macdonald et al., 1996). Such fault widening would increase crustal porosity and lead to higher conductivity in ridge-parallel planes. Elevated MgO concentrations observed in the subparallel damage and cataclastic zones of Hess Deep Rift crustal rocks imply ongoing fluid recharge through these ∼1 Ma samples (Hayman & Karson, 2007).

Heat flow measurements also point to basement outcrop-facilitated hydrothermal circulation in older, 18–22 Ma EPR-derived Cocos crust, which indicates high crustal permeability (∼10⁻¹⁰–10⁻⁹ m²) can persist and carry fluid below impermeable sediments (Fisher et al., 2003; Hutnak et al., 2008). Additionally, geodynamic models suggest that ∼10, ∼20, and ∼30 km deep vertical cracks in 10, 50, and 100 Ma lithosphere, respectively, can open because of thermal stress accumulation in cooling lithosphere (Korenaga, 2007). While such models refer to cracking in the upper lithospheric mantle, any thermal cracking that affects the lithospheric mantle should also affect the crust because the two layers are mechanically coupled. Though these cracks are likely sealed by serpentinization in the lithospheric mantle (Korenaga, 2017), the crust can support larger porosities. Thus, thermal-induced cracking should permit seawater to infiltrate porous vertical planes in the crust, increasing crustal conductivity in the paleo-ridge-parallel direction. The crustal anisotropy required by our data therefore reveals the influence of shallow lithospheric cracking and subsequent hydration as the plate matures and cools.

A seismic refraction study found crack anisotropy in oceanic lithosphere, but only as deep as crustal layer 2c, the sheeted dike complex (Shearer & Orcutt, 1985). Comparing inversions of active source seismic travel time tomography data to those of CSEM data collected in areas of hydrated bending faults at the Middle America Trench off Nicaragua reveals the ability of electrical data to image fluid-filled cracks at a greater resolution than these seismic data (Ivandic et al., 2008; Naif et al., 2015). The water-rich faults imaged there region. This new model has an improved RMS misfit of ∼1.02 and allows us to account for the second-order asymmetries, but at the expense of model parsimony.

4. Fluid Pathways and Sheared Olivine

Figure 14 summarizes our findings and interpretations.

4.1. Crustal Anisotropy

The crustal anisotropy of high planar conductivity relative to the plane normal direction is consistent with conductive, vertical sheets within an insulating, resistive matrix (Everett & Constable, 1999). The sheeted dike fabric of the upper crust should not necessarily generate this anisotropy unless a resistivity contrast exists between each dike. Additionally, the underlying gabbros, which exhibit anisotropy with the same direction of maximum conductivity are not emplaced in subvertical sheets (VanTongeren et al., 2015). We thus propose the crustal anisotropy indicates that cracks have created damage zones that act as fluid pathways, permitting hydration throughout the full depth extent of the crust. Though the 18–36 times magnitude of anisotropy sound like large values, parallel and series conduction models for bulk media can be used to show that they are merely equivalent to 4 and 0.8 mm wide, seawater-filled cracks spaced by 1 m in the upper and lower crust, respectively.

Although there is evidence of abyssal hill faulting in the area, global bathymetric observations generally do not show deep cutting normal faults on fast-spreading-derived crust (Carbotte & Macdonald, 1994). To address this inconsistency, we propose that the observed crustal anisotropy began to form at the mid-ocean ridge through normal faulting and hydrothermal microcracking (Manning et al., 2000), which created ridge-parallel zones of weakness. Such weaknesses promoted deep cracking and openings that allowed for seawater infiltration as the plate aged.
are 5 times more conductive in the fault plane than the plane perpen- dicular direction (Key et al., 2012; Figure 6). Perhaps electrical anisotropy is similarly more sensitive to hydration than seismic anisotropy.

While we do not expect the initial depth of abyssal hill normal faults to prevent thermal cracking of the entire crust, deeper faults at the ridge axis may facilitate off-axis cracking more readily than shallower faults. Therefore, in areas of comparably thin sediment cover, we expect slow-spreadig derived crust with deeper initial faults to be more anisotropic than the fast-spreadig derived crust of the APPLE survey region.

4.1.1. Sensitivity to Crack Depth

Cracks are not thought to penetrate deeply into the crustal section of fast-spreadig-derived lithosphere (Carbotte & Macdonald, 1994). To gain a better understanding of how well our data can constrain the depth extent of crustal anisotropy and the inferred cracks, we forward modeled the response of TIY crustal anisotropy for various crack depths (Figure 15). These models suggest that the characteristic $P_{\text{max}}$ crustal dimple only develops for anisotropy, and hence cracking, which penetrates both the upper and lower crust (see 6 km thickness $P_{\text{max}}$ response in Figure 15).

The model studies of sections 3.2.1 and 3.2.2 indicate that the $P_{\text{max}}$ dimple might also form with a more anisotropic or an overall more conductive upper crust atop an isotropic lower crust. To test if the real data are compatible with an isotropic lower crust, we reinverted the data, forcing isotropy in the gabbros via a 1,000 times higher penalty against anisotropy in the lower crustal layer. The resulting model (not shown) has a more anisotropic dike layer (37 times compared with 18 times in the preferred model) with lower bulk resistivity, but a higher RMS misfit (1.41 compared to 1.36 in the preferred model). An F test allows us to reject the null hypothesis, that the difference in misfits is due to random variation, with 70% confidence. So our data require anisotropy in both the dikes and gabbros.

Seismic data would be much less sensitive to pore water in gabbros than EM data, so it is unsurprising that full crustal cracking may have eluded seismic studies on intraplate lithosphere. Our finding of fluidized cracks through the entire crust is therefore a unique contribution to our understanding of oceanic lower crust and its potential hydration.

4.2. Uppermost Mantle Anisotropy

In the uppermost mantle, the paleo-spreadig direction is ~29 times more conductive than the vertical plane parallel to the paleo-ridge, a pattern that can be generated by horizontally aligned conductive rods embedded in an insulating matrix (Section 1.1; Everett & Constable, 1999). In contrast to the crust, we argue that the uppermost mantle inherits electrical anisotropy during its genesis near the ridge axis.

This mantle anisotropy cannot be explained by the intrinsic anisotropy of water-poor olivine since single crystals show anisotropy ratios of at most ~3 at uppermost mantle temperatures (Duba, 1972). Neither can hydrous olivine be the source since it only exhibits large anisotropy for water contents that exceed olivine’s storage capacity at upper mantle conditions (Gardes et al., 2014; Mierdel et al., 2007).

A possible contributor to this anisotropy is the interconnection of a conductive mineral phase (Everett & Constable, 1999; Watson et al., 2010). Conductive sulfide minerals have been found along grain boundaries and fractures in peridotitic xenoliths (Ducea & Park, 2000). Graphite ($\sim 10^{-5} \, \Omega \text{m}$) or solid carbon, if abundant (>100 ppm), stable, and interconnected, could explain high conductivities in the mantle (Duba & Shankland, 1982), though recent studies have called into question the thermodynamic stability of interconnected graphite on olivine grain boundaries at upper lithospheric mantle conditions (Zhang & Yoshino, 2017). Regardless of stability, any conductive mineral phase would require a mechanism for alignment in the paleo-spreadig direction to generate the modeled anisotropy.

On the other hand, recent experimental results on sheared San Carlos olivine aggregates measured electrical anisotropy ratios greater than 10 due to grain boundary alignment (Pommier et al., 2018), suggesting that mantle shearing could generate our observed anisotropy. Direct extrapolation of these data for $T < 850^\circ C$ and a shear strain of $\sim 1.3$ yields resistivities for the shear-parallel and shear-plane normal directions that are
in excellent agreement with our model at a temperature of 600°C (Pommier et al., 2018; Figure 6). However, the uppermost mantle temperature of 33 Ma lithosphere is expected to be about 330°C (Stein & Stein, 1992), where the sheared olivine data predict much higher resistivities ($\rho_{\text{shear} \parallel} = 6 \times 10^4 \, \Omega \text{m}$; $\rho_{\text{shear} \perp} = 5 \times 10^7 \, \Omega \text{m}$). While these experimental data explain the large anisotropy in our preferred model, they do not explain its absolute resistivities. There are likely other factors affecting the bulk resistivity of sheared mantle that have not yet been measured in laboratory studies. Inclusion of additional minerals such as water-bearing aluminous orthopyroxene, may help to account for our observed lower bulk resistivity (Dai & Karato, 2009; Mierdel et al., 2007), though hydrous pyroxene exhibits weak anisotropy (Yang et al., 2012). Further, the shear-induced alignment of grain boundaries, posited as generating the large anisotropies, likely depends on grain size, a parameter that was not extensively probed. Sheared samples in the study had grain sizes ranging from ~3–7 μm (Pommier et al., 2018), whereas uppermost lithospheric mantle grain sizes are expected to be on the order of millimeters (Turner et al., 2015). Additional lab studies are needed to characterize how shear-induced anisotropy varies with grain size, composition, and degree of shearing. Even so, the remarkable similarity of our preferred model’s uppermost mantle anisotropy with lab results on sheared olivine strongly suggests that shearing plays a dominant role in creating the electrical character of the uppermost mantle and may define the electrical response of the Moho.

Alignment of olivine by mantle flow is widely held as the mechanism by which oceanic lithospheric mantle inherits seismic anisotropy (Gaherty et al., 2004; Kodaira et al., 2014; Mark et al., 2019; Russell et al., 2018; Shearer & Orcutt, 1985; Vanderbeek & Toomey, 2017). Such seismic anisotropy studies have determined that the fast $a$ axis of olivine is aligned with the flow direction in the uppermost mantle (Karato et al., 2008), although the degree of this anisotropy is variable and seems to correlate with spreading rate. Gaherty et al. (2004) estimated a 3.1–3.7% $P$ wave anisotropy in uppermost mantle of the slow-spreading-derived western Atlantic. Vanderbeek and Toomey (2017), analyzing data from the intermediate spreading-derived Juan de Fuca Plate, measured a moderate $Pn$ azimuthal anisotropy of 4.2–5.0%. They suggested that the seismic anisotropy frozen into the uppermost mantle is in fact a combination of the original relative plate motion and the absolute plate motion directions because of the time it takes the lithospheric mantle to cool. A strong $P$ wave azimuthal anisotropy of 8.5–9.8% found in the uppermost mantle on fast-spreading-derived lithosphere was interpreted as having originated from drag force by mantle flow, which affected a thin (~2 km) layer directly beneath the Moho (Kodaira et al., 2014).

The APPLE survey lies on fast spreading (13 cm/yr; Müller et al., 2008) EPR-derived lithosphere and should, therefore, have a strong seismic anisotropy. But while the pattern of electrical anisotropy in our preferred model matches the direction of seismic anisotropy, alignment of the olivine crystal axes alone cannot explain electrical anisotropy as it can with seismic anisotropy, as the grain interior conductivity of olivine is neither high enough at the expected temperatures nor has a large enough anisotropy. At present, the authors are unaware of any study that considers how the electrical anisotropy of dry olivine varies with the amount of shear strain. Nevertheless, we suspect some minimum strain, and thus spreading, rate must be exceeded in order to generate anisotropy. Because large strain rates should occur within ~20 km of the mid-ocean ridge axis (Turner et al., 2017), we conclude that the observed electrical anisotropy arose from a shear-induced grain boundary fabric that froze into the shallow mantle early during plate formation.

4.3. Comparison to Anisotropy on the Cocos Plate Offshore Nicaragua

The only other section of oceanic lithosphere where CSEM data have been collected using circular tows is the Cocos Plate at the Middle America Trench off Nicaragua (Key et al., 2012). There, two 30 km radius tows were completed in 2010. One of the circular tows, located on the abyssal plain about 130 km seaward of the trench, was meant to provide a reference for the anisotropy of the incoming Cocos plate. The other circle covered the trench outer rise wall where plate bending has led to pervasive normal faulting. Analysis of these data in Key et al. (2012) focused on comparing the highly anisotropic $P_{\text{max}}$ signal seen at the trench outer rise with the relatively isotropic-looking $P_{\text{max}}$ data on the abyssal plain. We now present a closer look at the abyssal plain data after taking 2,400 s long data stacks, compared to the 120 s stacks used in Key et al. (2012), to reduce the variance in $P_{\text{max}}$ and $P_{\text{min}}$.

The newly stacked data are shown in Figure 16 for the three strongest transmitter frequency harmonics. In all frequencies, the $P_{\text{min}}$ data contain a small cloverleaf pattern that resembles the anisotropic patterns observed in the APPLE $P_{\text{min}}$ data and model studies described in Section 3.2. Due to the thicker sediments offshore Nicaragua (~500 m), these data have much lower amplitudes than the APPLE data,
and a direct comparison between the two studies is further complicated by the different transmission frequencies. However, the existence of the $P_{\text{min}}$ cloverleaf with lobes elongated between the orthogonal paleo-ridge-parallel and paleo-spreading directions suggests the presence of anisotropy on the abyssal plain of the Cocos plate. Though further modeling and inversion are needed to quantify this anisotropy, the observation of azimuthal electrical anisotropy in two separate oceanic plates where suitable CSEM data have been collected begs the question of whether other oceanic plates have similar electrical anisotropy.

5. Conclusions

Our observations of azimuthal electrical anisotropy on the Pacific Plate provide new windows into crustal hydration and mantle deformation. The crustal anisotropy observed in the APPLE data suggests that abyssal hill normal faulting, which begins at the ridge axis, matures with thermal contraction-induced cracking and allows hydration to affect the entire crust. The uppermost mantle anisotropy points to a frozen-in, paleo-flow fabric that developed from shearing close to the ridge axis. Anisotropy also seen on the Cocos Plate suggests that this newly discovered electrical anisotropy may be prevalent in other oceanic plates. Additional CSEM anisotropy studies are needed to determine whether such electrical anisotropy is a universal feature of oceanic lithosphere, and to investigate how crustal hydration and mantle paleo-flow vary with spreading rate and across discontinuities, such as transform faults.

References


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